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Southern Patagonian glacial chronology for the Last Glacial period and implications for Southern Ocean climate

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Abstract

The Magellan region of southern South America is in a unique setting, at $>50°S$ on the equatorial side of the Antarctic Frontal Zone, to record in detail terrestrial glacial to interglacial events. A ^{10}Be chronology shows growth and millennial fluctuations of a Patagonian Ice Sheet between ~25 and 17.6–17.0 cal ka. In the Strait of Magellan, the maximum ice margin position is dated to 24.6 ± 0.9 ka, and other moraine ages are 18.5 ± 1.8 and 17.6 ± 0.2 ka (mean ± 1 standard deviation). In Bahía Inútil, dated moraine ages are 20.4 ± 1.2 and 17.3 ± 0.8 ka. The chronology of the local Last Glacial Maximum (LGM) reveals a record of atmospheric cooling that was broadly in phase with changes in Southern Ocean conditions, such as sea-ice fluctuations and surface water characteristics. Published modeling results indicate that a decline in temperature of ~6 °C and slight drying over southernmost Patagonia could simulate the growth and sustained presence of an ice sheet to the mapped LGM limit. The terrestrial record in southern Patagonia and marine records in adjacent oceans indicate mean northward movement of the Antarctic Frontal Zone, which caused the last southern South American ice age. The Antarctic Frontal Zone at present lies only 3–5° to the south. Some significant changes in the Magellan region occurred in step with North Atlantic region and the Northern Hemisphere. For example, the overall time span of the last glaciation and the timing of maximum ice extent were similar between the hemispheres, despite maximum local summer insolation intensity in southern South America. Other characteristics of the southern Patagonian glacial history differ from the North Atlantic region, specifically an out-of-phase relationship during deglaciation, which is more similar to that of Southern Ocean and Antarctic records.

1. Introduction

Resolving the timing and structure of the Last Glacial period around the globe is important because such knowledge can help evaluate whether the tropics, North Atlantic Ocean, or Southern Ocean may drive abrupt regional and global climate changes. In particular, the Southern Ocean, dominated by the Antarctic Circumpolar Current, affects all three major oceans, influences deep water worldwide, is one of the Earth's major heat sinks, and is an important control on atmospheric CO_2 (Toggweiler and Samuels, 1995; Francois et al., 1997; Knorr and Lohmann, 2003). A lack of terrestrial data hinders understanding of the southern latitude air–ocean system and its role in climate dynamics during the global Last Glacial Maximum (LGM) (e.g., 23–19 ka, Mix et al., 2001). Marine records, although invaluable, are less common and are susceptible to ^{14}C reservoir effects (Charles et al., 1996; Lamy et al., 2004). Also, they are only an indirect proxy for past atmospheric conditions, and do not characterize spatial–temporal terrestrial climate variability. Moreover, there are few land areas in middle to high southern latitudes and long ice core records are limited to Antarctica.

A terrestrial record spanning the local LGM near the northern boundary of the Antarctic Frontal Zone, where

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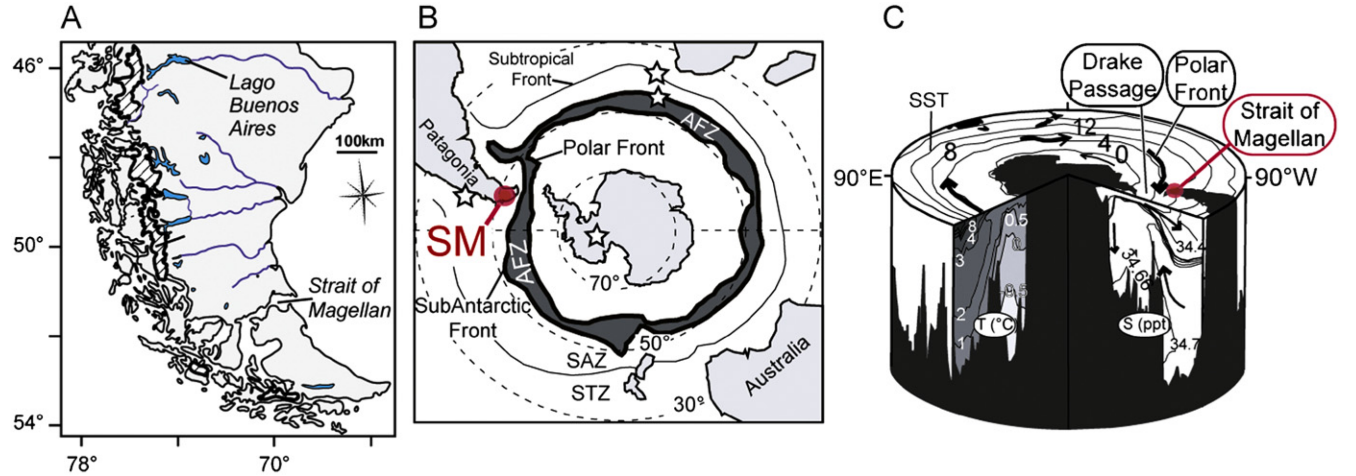


Fig. 1. Southernmost South America intruding into the mid-high latitude southern ocean. (A) Location of the Strait of Magellan (SM) area. (B) Map of present Antarctic Frontal Zone (AFZ), Sub-Antarctic Zone (SAZ), and Subtropical Zone (STZ) (Belkin and Gordon, 1996). The Magellanes area is within the SAZ, which has an annual temperature of ~6–7 °C. Detailed characteristics of each climate zone presented in Belkin and Gordon (1996); the Polar Front, SubAntarctic Front, and Subtropical Fronts are roughly associated with the ~1–2 °C, 5–10 °C, and >10 °C isotherms, respectively. Marine and the Byrd Antarctic ice core records in Fig. 4 are shown with stars. (C). Cross section west of the Drake Passage (from Olbers et al. (2004)), showing surface and subsurface temperatures and salinity, core of eastward flowing Antarctic Circumpolar Current and associated polar front near the 4 °C isotherm (black arrow on surface), upwelling of poleward flowing saline deepwater (up arrow), which is balanced in part by low salinity northward flowing Antarctic immediate water (arrow within 34.4), and sinking of cold slope water (downward arrow).

temperature drops considerably by 6 °C or more (Fig. 1b, Belkin and Gordon, 1996; Olbers et al., 2004), is a proxy for former changes in atmospheric conditions and allows linking with middle to high-latitude marine and ice core records. In this paper, we present a glacial chronology during the LGM from just north of the Drake Passage, on the most southerly continental setting outside Antarctica (Fig. 1), <3–5° north of the 'modern interglacial' position of the Antarctic Frontal Zone. Southernmost South America is less than 1000 km from the Weddell Sea where deepwater formation occurs (Fig. 1b) and is the only continental land mass between ~45 and 65°S. Glacial records are one of the best proxies of past atmospheric behavior and snowline change outside the ice-covered regions, and those from southern South America are specifically proxies of the former behavior of the westerlies and air–ocean systems surrounding Antarctica (Fig. 1).

A firm LGM chronology for the southernmost Andes has been lacking largely because of a paucity of material for ^{14}C dating, especially in arid environments such as in the lee of the Andes. The west side of the southernmost Andes was covered by LGM ice and thus the glacial geologic record is primarily one of deglaciation. In addition, available proxy paleoclimate records from middle and low latitude South America have been used to argue for in-phase and out-of-phase behavior with records in both the Southern Ocean and the North Atlantic region during the Last Glacial period and interglacial transition (Denton et al., 1999; Bennett et al., 2000; Gilli et al., 2001; Heusser, 2003; Kaplan et al., 2004; Smith et al., 2005; Sugden, 2005).

We measured in situ cosmogenic ^{10}Be accumulated in erratics that closely date glacial landforms recording former ice margin positions of Andean ice lobes in the Strait of Magellan and Bahía Inútil, ~53–54°S (Fig. 2). This investigation ties together new and previous data (e.g., Heusser, 2003; McCulloch et al., 2005), which, collectively, allows a better comparison between the cosmogenic nuclide chronology of Patagonian glacial events and other records of southern latitude climate change, including the adjacent Southern Ocean, during the last major glacial period and termination.

2. Setting

The Strait of Magellan and Bahía Inútil are major marine inlets surrounded by low-lying coastal areas within the southern tip of Patagonia, extending towards the dry steppe of semi arid eastern Patagonia (Fig. 2). Low-gradient outlet glaciers in the southern part of the ice sheet extended northeast to east down the Strait of Magellan and Bahía Inútil towards the South Atlantic Ocean, where glacial landforms that indicate major ice margin stillstands or readvances have been mapped in detail (Fig. 2) (Porter, 1990; Clapperton et al., 1995; Benn and Clapperton, 2000; Rabassa et al., 2000; Bentley et al., 2005). At the maximum of the Last Glacial period, total Patagonian ice sheet volume was ~500,000 km^3, the equivalent of ~1.2 m of global sea level (Hulton et al., 2002). A moraine complex on Peninsula Juan Mazia marks the greatest LGM extent of Magellan ice. Glaciotectonic evidence shows that, at least on parts of the peninsula, the moraines represent ice margin readvances (Benn and Clapperton, 2000). The distance from the LGM divide to ice margin was ~350–400 km and surface slopes and driving stresses are estimated to be as low as ~7.5 m/km and <10 kPa,

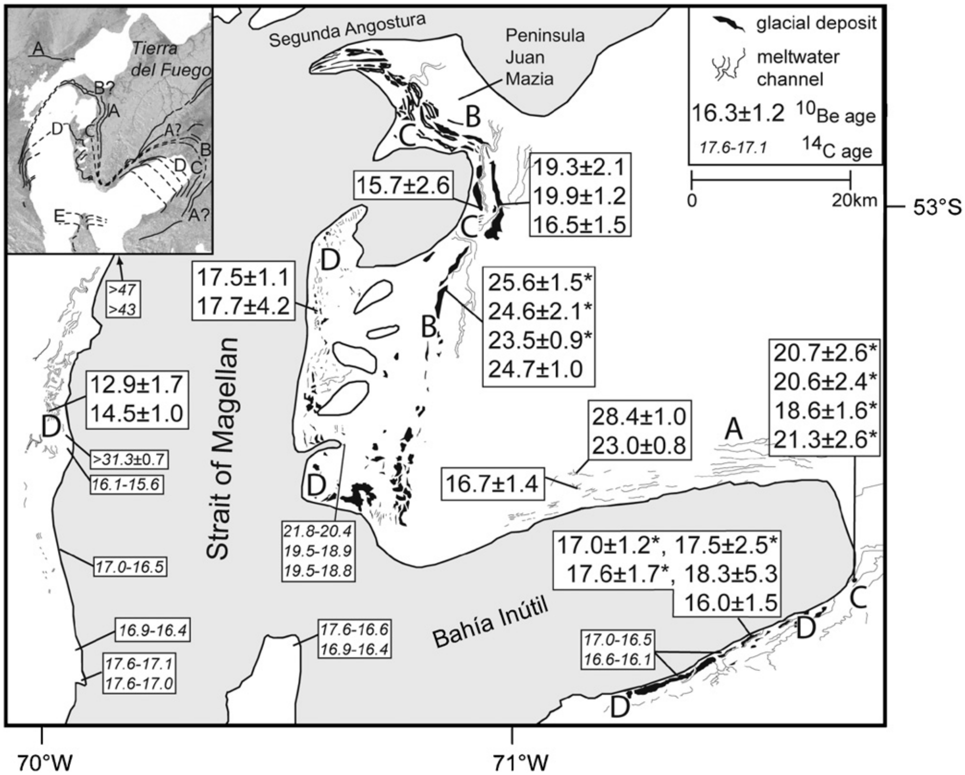


Fig. 2. Simplified map of glacial deposits and meltwater channels around the Strait of Magellan and Bahía Inútil based on Clapperton et al. (1995), Benn and Clapperton (2000) and Bentley et al. (2005), along with cosmogenic surface exposure ages (ka, erosion rate = 0 mm/ka) from this study and McCulloch et al. (2005; the ages with asterisks). Inset shows major moraine limits of the last glaciation in the Strait of Magellan and Bahía Inútil; the A limit is pre-LGM and E limit the inferred location of Lateglacial ice (from Bentley et al., 2005). Large letters in caps near mapped glacial deposits correspond to those in the inset. Selected, key ^{14}C ages from cores and stratigraphic sections are shown (1σ calibrated age range; Heusser, 2003; McCulloch et al., 2005). ^{14}C ages on the west side of the Strait of >31 ka, and often infinite in age, bear on the timing of the start of the LGM. Ages on the east side of the Strait of ~22–19 constrain onset of deposition after retreat from the C limit, and ^{14}C ages in both marine embayments firmly constrain the timing of deglaciation of the entire area by ~17.6–17 ka (Fig. 3).

respectively (Clapperton et al., 1995; Benn and Clapperton, 2000). Such low surface gradients and driving stresses imply that the ice lobes would have had a relatively rapid response to abrupt snowline change.

The source (accumulation) region of ice for the glacial geologic record presented here is located between the Antarctic Frontal Zone and the core of the middle latitude westerlies (~50°S). Such a location should act as a '3-D dipstick' recording past glacial to interglacial movement of the mid to high-latitude air–ocean masses (Prohaska, 1976) that surround Antarctica (Fig. 1). Presently, the westerlies nourish glaciers in the Andes in this latitudinal range. Modern temperature and steep precipitation gradients across the southern Andes are due to topography and orographic effects. Precipitation varies from ~2000 mm/yr along the western coast to <500 mm/a in the area studied (Fig. 2) (Prohaska, 1976), where annual temperature is about 6–7 °C.

3. Surface exposure dating

Sampling strategies aimed to refine the overall chronology of McCulloch et al. (2005) and the mapping of Bentley et al. (2005) and Benn and Clapperton (2000), which were guides to collecting. They defined at least three major moraine 'limits', 'B–D,' for the Last Glacial period along Bahía Inútil and the Strait of Magellan, and one Lateglacial event, E, confined to the mountains at the head of the Strait (Fig. 2). In this study, 14 additional dates supplement 10 previously recorded (Table 1; McCulloch et al., 2005). For the Magellan outlet lobe, we measured samples collected from five glacial moraines (Fig. 2). Two moraines are on the east side of Peninsula Juan Mazia (mapped limit B–C; SM-02-09, 11, 12, 14), and one additional sample (limit B; BGG:C3) was measured from a previously dated deposit just to the southeast. Two moraines are on the east and west sides of the Strait of Magellan, respectively (limit D; SM-02-01, 04, 07, 08). For the Bahía Inútil lobe, four additional ages were obtained. Two ages are from a previously dated moraine on the south side (limit D; SM-02-21, 23). On the north side of the bay, two ages were measured on samples (SM-02-17 and 18) above a distinct yellowish 'boundary' around ~325 m, demarcating thicker loess cover above the A–B limit (Bentley et al., 2005). One new age is on an erratic (SM-02-20) just below the A–B limit.

We preferentially selected the largest boulders on, or nearly on, exposed moraine crests, which are assumed to be

Table 1
Sample location and information

Sample	Lat. °S	Long. °W	Elev (km)	Boulder height (cm)	Sample mass (g)	Sample thickness (cm)	Production rate (atoms/g/yr)
Strait of Magellan							
SE of Juan Mazia Peninsula (B limit)							
BGGC1*	53.07	70.15	0.082	100	38.612	5.9	5.84
BGGC2*	53.08	70.14	0.100	200	52.002	5.9	5.95
BGGC3	53.08	70.14	0.100	100	15.17	3.0	6.00
BGGC4*	53.08	70.15	0.073	200	22.977	3.9	5.79
Juan Mazia Peninsula, outermost moraine (C limit)							
SM-02-09	52.92	70.04	0.049	35	8.822	0.9	5.80
SM-02-11	52.91	70.04	0.042	35	33.050	1.5	5.73
SM-02-12	52.90	70.04	0.045	163	13.634	3.0	5.68
Juan Mazia Peninsula innermost moraine (C limit)							
SM-02-14	52.93	70.06	0.037	30	6.058	0.5	5.75
Second youngest moraine dated (D limit)							
SM-02-07	53.07	70.44	0.031	90	26.241	0.5	5.72
SM-02-08	53.07	70.43	0.033	50	24.291	0.8	5.71
Youngest moraine dated (D limit)							
SM-02-01	53.26	71.00	0.134	25–30	27.446	0.5	6.33
SM-02-04	53.26	71.00	0.123	19–23	23.400	2.0	6.18
Bahia Inutil							
Glacial deposit at end of bay (B limit)							
BIB1*	53.50	69.25	0.080	800	76.358	0.9	6.02
BIB2*	53.50	69.23	0.075	900	76.358	0.0	5.84
BIB3*	53.51	69.25	0.085	700	33.618	5.0	5.90
BIB4*	53.51	69.25	0.080	500	57.291	3.9	5.87
Central bay (D limit)							
BIC1*	53.58	69.48	0.052	300	43.129	3.9	5.71
BIC2*	53.58	69.49	0.048	200	56.619	5.9	5.68
BIC3*	53.59	69.49	0.054	200	19.262	5.9	5.72
SM-02-21	53.58	69.49	0.060	200	10.999	0.6	5.91
SM-02-23	53.58	69.49	0.056	115	31.275	2.9	5.78
Lateral moraines, north side (A or B limit?)							
SM-02-17	53.30	69.89	0.352	105	27.295	0.5	7.78
SM-02-18	53.30	69.89	0.350	100	21.465	0.5	7.76
SM-02-20	53.32	69.89	0.320	74	27.596	0.9	7.57

Note: No topographic shielding correction or geomagnetic corrections (53°S) required. Procedural blanks: $^{10}Be = 1.8$ and 1.7×10^{-14} (SM-02 samples), 2.5 or 4.1×10^{-14} (BIB and BIC), and 2.8×10^{-14} (BGG). BIB, BIC, and BGG samples (with asterisks) are from McCulloch et al. (2005). All samples measured at ETH-Zurich except BGGC3, SM-02-17 and 18, which were measured at SUERC (blank = 1.9×10^{-14}). Scaling factors based on production rate of 4.98 atoms/gm/yr (at sea level and high latitude) (Gosse and Stone, 2001; Version 1 at http://hess.ess.washington.edu-/math/al_be_stable/al_be_multiple.php), atmospheric pressure of 1001.71 mbar (Strait of Magellan) or 1000.94 mbar (Bahía Inútil), and temperature of 281 K (Stone, 2000).

stable and to have escaped human disturbance and burial by snow or volcanic ash. Boulders with evidence of rock splitting or spallation were avoided. Using a hammer and chisel, samples were collected from the upper few cm of boulders, at least 5 cm from edges and sharp facets. Elevations were measured relative to sea level using a barometric-based altimeter, and cross-checked against GPS, and are estimated to be correct within 10 m. Snow cover is not corrected for but in view of the limited annual precipitation of <500 mm, with no appreciable snow accumulation, it is assumed to have had an insignificant effect on ages (Gosse and Phillips, 2001). Quartz was separated and purified at the University of Edinburgh following the method outlined in Bierman et al. (2002). $^{10}Be/^{9}Be$ isotope ratios were measured at the AMS facility in Zurich operated jointly by the Paul Scherrer Institut and ETH Zurich, or the Scottish Universities Environmental Research Centre (Table 1).

For all ^{10}Be age calculations, including measurements previously published, we use the ^{10}Be–^{26}Al exposure age calculator at http://hess.ess.washington.edu/math/al_be_stable/al_be_multiple.php (Version 1). This program uses a global production rate of 4.98 ± 0.34 atoms/g/yr for ^{10}Be ($\pm 2\sigma$; sea level and high-latitude and standard atmosphere), which was updated from 5.1 atoms/g/yr (Gosse and Stone, 2001; Stone, 2000). Production by neutron spallation and muons are 4.87 ± 0.33 and 0.11 ± 0.01 atoms/g/yr, respectively (Stone, 2000). Production rates are corrected for a local air pressure of ~1002 mbar (Table 1) and annual temperature of 281 K (Taljaard et al., 1969; Van Loon et al., 1972) and scaling to altitude and geographic latitude is based on Stone's (2000) reformulation of that in Lal (1991). Ages are presented without erosion. However, ages are also shown in Table 2 with an erosion rate of 1.4 mm/ka, which is based on that derived for southern Patagonia (Kaplan et al., 2005), to highlight that this effect is <3%. Because of the latitude, 53°, changes in geomagnetic field strength are assumed to cause <1% change in ages (Gosse and Phillips, 2001). Depth scaling follows exponential decay as described in Gosse and Phillips (2001) with rock density = 2.65 g/cm^3 and attenuation length 145 ± 7 g cm^2 (Brown et al., 1992).

Individual exposure ages are shown and discussed with analytical errors only ($\pm 1\sigma$; Table 2). Although a possible systematic uncertainty must be kept in mind when studying the cosmogenic ages presented here, we emphasize that such an uncertainty does not affect the main conclusions presented, especially given the existence of radiocarbon data for the deglacial period (Heusser, 2003; McCulloch et al., 2005). Furthermore, various schemes that scale production rates of cosmogenic nuclides to latitude and elevation agree within 5% for southern Patagonia (Dunai, 2000, 2001; Stone, 2000; Desilets and Zreda, 2003; Lifton et al., 2005; Desilets et al., 2006). Atmosphere pressure may have been different in the glacial world (Ackert et al., 2003), but this effect on cosmogenic ages presented here is inferred to be well within 10%, because most of the exposure history of the samples includes the pressure of the interglacial period (Stone, 2000).

Table 2
^{10}Be data in the Strait of Magellan and Bahía Inútil

Sample	^{10}Be (10^4 atoms/g)	Age1 No erosion ka ± 1σ	Age2 Erosion ka ± 1σ
Strait of Magellan			
SE of Juan Mazia Peninsula (B limit)			
BGGC1*	14.90 ± 0.85	25.6 ± 1.5	26.5 ± 1.5
BGGC2*	14.54 ± 1.24	24.6 ± 2.1	25.4 ± 2.2
BGGC3	14.72 ± 0.60	24.7 ± 1.0	25.5 ± 0.9
BGGC4*	13.54 ± 0.49	23.5 ± 0.8	24.2 ± 0.9
Average age = 24.6 ± 0.9 (weighted mean = 24.3 ± 0.6)			
Juan Mazia Peninsula, outermost moraine (C limit)			
SM-02-09	11.11 ± 1.19	19.3 ± 2.1	19.7 ± 2.1
SM-02-11	11.34 ± 0.68	19.9 ± 1.2	20.4 ± 1.2
SM-02-12	9.318 ± 0.84	16.5 ± 1.5	16.8 ± 1.5
Average age = 18.5 ± 1.8 (weighted mean = 18.7 ± 0.8)			
Juan Mazia Peninsula innermost moraine (C limit)			
SM-02-14	8.993 ± 1.51	15.7 ± 2.6	16.0 ± 2.7
Second youngest moraine dated (D limit)			
SM-02-07	9.942 ± 0.64	17.5 ± 1.1	17.9 ± 1.1
SM-02-08	10.07 ± 2.40	17.7 ± 4.2	18.1 ± 4.3
Average age = 17.6 ± 0.2 (weighted mean = 17.5 ± 1.1)			
Youngest moraine dated (D limit)			
SM-02-01	8.166 ± 1.06	12.9 ± 1.7	13.2 ± 1.7
SM-02-04	8.908 ± 0.60	14.5 ± 1.0	14.7 ± 1.0
Average age = 13.7 ± 1.1 (weighted mean = 14.1 ± 0.08)			
Bahía Inútil			
Glacial deposit at end of bay (B limit)			
BIB1*	12.79 ± 1.55	21.3 ± 2.6	22.0 ± 2.7
BIB2*	11.99 ± 1.39	20.6 ± 2.4	21.2 ± 2.5
BIB3*	10.95 ± 0.96	18.6 ± 1.6	19.1 ± 1.7
BIB4*	12.10 ± 1.52	20.7 ± 2.6	21.3 ± 2.7
Average age = 20.4 ± 1.2 (weighted mean = 19.9 ± 1.1)			
Central bay (D limit)			
BIC1*	9.645 ± 0.68	17.0 ± 1.2	17.3 ± 1.2
BIC2*	9.885 ± 1.42	17.5 ± 2.5	17.9 ± 2.6
BIC3*	10.00 ± 0.98	17.6 ± 1.7	18.0 ± 1.8
SM-02-21	10.80 ± 3.14	18.3 ± 5.3	18.8 ± 5.5
SM-02-23	9.240 ± 0.86	16.0 ± 1.5	16.4 ± 1.5
Average age = 17.3 ± 0.8 (weighted mean = 16.9 ± 0.8)			
Lateral moraines, north side (A or B limit?)			
SM-02-17	21.97 ± 0.75	28.4 ± 1.0	29.5 ± 1.0
SM-02-18	17.78 ± 0.60	23.0 ± 0.8	23.4 ± 0.8
SM-02-20	12.59 ± 1.05	16.7 ± 1.4	17.1 ± 1.4

Note: ^{10}Be dates associated with glacial deposits in the Strait of Magellan and Bahía Inútil. Moraine limits from Bentley et al. (2005). Ages with asterisks from McCulloch et al. (2005). Age1 is without erosion. Age2 is with an erosion rate of 1.4 mm/ka. For a glacial landform, arithmetic average calculated (with Age 1) only if at least two ^{10}Be ages exist for the deposit, and the error is the standard deviation of individual boulder ages. All ages calculated with http://hess.ess.washington.edu-/math/al_be_stable/al_be_multiple.php, Version 1. 1σ error includes analytical or measurement uncertainty, including that for the procedural blank.

4. Prior ^{14}C dating and amino acid data

Radiocarbon and amino acid racimization data on marine shells in the Strait of Magellan provide additional chronometers (cf., ^{10}Be ages) to reconstruct the glacial history, especially at the beginning and end of the LGM (McCulloch et al., 2005). On the western shore of the Strait of Magellan and the eastern shore of Isla Dawson, ^{14}C ages on reworked shell fragments, from diamict inside limit B, range from >40,000 (i.e. infinite) to 27,690 ^{14}C yr BP (44,760 ± 460 to 31,250 ± 670 cal yr BP). These ages reflect an ice-free period when the Strait is connected to the sea, before the LGM (McCulloch et al., 2005). During the LGM, when eustatic sea level was lower than today (e.g., at least 120 m between ~23 and 19 ka), the area north of Segunda Angostura was a land barrier between the Atlantic Ocean and the study area, as glacial ice blocked drainage to the Pacific Ocean. Based on amino acid analyses Clapperton et al. (1995) estimated ages

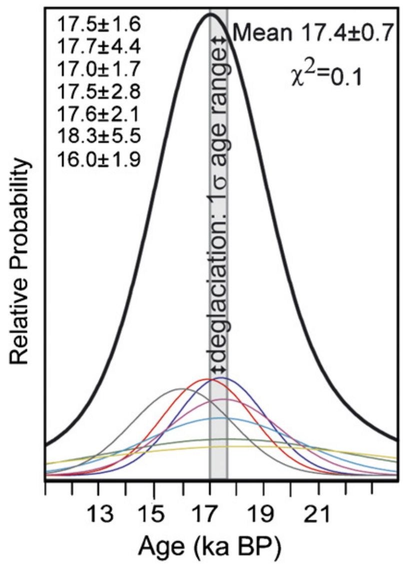


Fig. 3. The last major moraine-building event before deglaciation of Magellan Strait and Bahía Inútil. Ages from the *D* limits are shown (Table 2, erosion rate = 0 mm/ka) in a relative probability plot. Colored lines are individual ages and black line represents the probability distribution for the dataset. Also shown is the calibrated 1σ age range for deglaciation based on the oldest ^{14}C ages, ~17.6–17 ka, on Fig. 2. Reduced chi-squared (χ^2) values of <1 indicate that the cosmogenically dated samples are from the 'same' population and that the analytical errors alone could explain the age scatter (Bevington and Robinson, 1992). Given the uncertainties in both chronometers the data are consistent. Taken at face value, they imply that the last prominent moraine-building event was followed rapidly by deglaciation of the Strait, i.e., within the resolution of the dating tools.

>35,000–90,000 yr BP for shell samples around the Strait, which were partly calibrated by finite radiocarbon dates on fossil fragments in the same tills. The lack of paleotemperature data prevents confident assignment of ages using amino acid data, but nonetheless, the measurements are useful for constraining the timing of ice-free conditions before the LGM (Clapperton et al., 1995).

Numerous ^{14}C dates constrain the timing of deglaciation of both the Strait of Magellan and Bahía Inútil, which are summarized in Heusser (2003) and McCulloch et al. (2005). Most recently, McCulloch et al. (2005) carried out a ^{14}C dating campaign and revised the chronology for deglaciation. Several ^{14}C ages (calibrated) are >17 and 17–16 ka for deglaciation of the Strait of Magellan and Bahía Inútil. These ages are from stratigraphic sections and the bottom of lake cores. Only the oldest, >16 ka, are shown on Fig. 2. The 1σ calibrated age range, ~17.6–17.0 ka, for the oldest samples from Table 5 of McCulloch et al. (2005) is shown in Fig. 3. Subsequently, the evidence indicates glacial ice did not exist past the inner Strait (Heusser, 2003; McCulloch et al., 2005).

5. Results

Twenty-four ^{10}Be ages on glacial erratics constrain the timing of former ice margin fluctuations in the Magellan region (Fig. 2; Table 1). All ages shown and discussed assume an erosion rate of 0 mm/ka, unless otherwise mentioned. Dates from the B–C limits on Peninsula Juan Mazia range from ca 25.6±1.5 to 15.7±2.6 ka and suggest limited overall net retreat of the ice margin during this time period. The four boulders from the oldest landform provide the ages of ~26–24 ka (mean and SD of ~24.6±0.9 ka). The innermost (i.e., youngest) D moraine along the eastern side of the Strait yielded two ages of 17.7 and 17.5 ka (Fig. 2). Along the west side of the Strait, two boulders in close proximity to each other on a moraine crest produced ages of 14.5 and 12.9 ka.

For the Bahía Inútil lobe, the terminal ice margin extent (B limit) during the Last Glacial period lies just beyond the marine embayment (Clapperton et al., 1995; Rabassa et al., 2000) and remains undated (Mcculloch et al., 2005). Prominent 'C and D' ice margin positions are dated along the south side of the marine embayment (Fig. 2). Four ages on the 'C' landform range from ~21 to 19 ka (mean 20.4±1.2 ka; McCulloch et al., 2005), and five ages on the 'D' landform range from ~18 to 16 ka (mean ~17.3±0.8 ka). Regarding the latter, it is important to note that the ages are not from the outermost mapped D limit (Fig. 2). On the north side of Bahía Inútil, the two ages above the A–B 'boundary' (Bentley et al., 2005) are 28.5±1.0 and 23.1±0.8 ka, and one age at a lower elevation than the thick loess limit at ~325 m is 16.7±1.4 ka.

6. Discussion

6.1. Chronology of the Last Glacial period

Building upon prior studies, the chronology presented here more firmly constrains the glacial history of southern Patagonia and necessitates slight refinement in the ages and mapping of the moraine limits B–D as recognized by Clapperton et al. (1995) and McCulloch et al. (2005). The new ages suggest that much of the Peninsula Juan Mazia deposits are statistically younger than the ~25 ka 'B' landform just to the southeast as they have a mean age of 18.5±1.8 (Fig. 2; Table 2). The deposits that were dated along (at least) the eastern Peninsula Juan Mazia appear to be C in age; the B limit dated just to the southeast was preserved during the later C advance, perhaps because it escaped frontal outwash erosion. The Peninsula Juan Mazia ages also overlap within error with the outer moraine ages in Bahía Inútil, 20.4±1.2. Two new ages for a 'D' limit moraine on the east side of the Strait of Magellan and two additional ages on the southeast side of Bahía Inútil suggest coeval moraine-building events in both embayments at 17.6±0.2 and 17.3±0.8, respectively. Two ages of 28.5 and 23 ka from the upper lateral moraines along the north side of Bahía Inútil are from an area inferred to be pre-LGM in age, given thick loess cover, weathering development, and correlation to other areas (Bentley et al., 2005). Given there are only these two ^{10}Be ages, it is not possible to conclude whether mapping refinements are needed for this particular area, or these boulders were deposited prior to the LGM and have been affected by geomorphic processes such as intense erosion, as evident on pre-LGM glacial deposits to the east of the area studied (Kaplan et al., 2007).

The ^{10}Be chronology agrees within 1σ with ^{14}C (given the uncertainties in both chronometers) stratigraphic, and pollen data in the Strait of Magellan and Bahía Inútil area (Heusser, 2003; McCulloch et al., 2005). ^{14}C and racimization data suggest that following a relatively long ice-free period, the last glaciation began after ~31 ka. Maximum ice extent was around 25 ka and overall or net ice recession was quite limited until 19 ka. Given the uncertainties with both chronometers, three ^{14}C ages on the eastern side of the Strait (Fig. 2) of ~22–19 ka, marking deposition after retreat from the 'C' limit, are consistent at 1σ with the ^{10}Be ages on these deposits. The ^{10}Be chronology indicates that the last major moraine-building event occurred between ~18 and 17 ka.

To examine further the end of the LGM the ages from the dated D limits are combined on Fig. 3. Stratigraphic and ^{14}C data firmly indicate that ice never extended past the innermost Strait after ~17.6–17.0 cal ka (1σ calibrated age range) (Fig. 2). Taken at face value, the ^{14}C and cosmogenic nuclide data collectively suggest that the last major glacial event was followed by rapid warming and deglaciation. Caution is warranted when assigning the exact age of the last major moraine-building event because of present uncertainties of production rates (Table 2). Scaling factors for adjusting reference production rates to the sample locations vary up to ~3.5% (Dunai, 2000, 2001; Stone, 2000; Desilets and Zreda, 2003; Lifton et al., 2005; Desilets et al., 2006). However, even an adjustment of 4% would support the inference of a glacial event quickly followed by deglaciation of the Strait. An erosion rate of 1.4 mm/ka (Table 2), if correct, would raise the ages by about 2–3%, and generate D moraine ages slightly older, but still overlapping in time (at 1σ) with the ^{14}C constraint. A major change in pressure over the entire history of each sample, compared to the last 60 years of instrumental record, could significantly change the ages. If the pressure in the glacial world was lower and the production rate higher (Ackert et al., 2003), the ages would decrease. However, the 'interglacial pressure,' is applicable for much or most of the exposure duration of the samples. Assuming fundamental changes in pressure have not occurred since deglaciation, the production rate variability, at least for the 18–17 ka erratics, due to this effect also should be $<$5%. The systematic uncertainties and erosion rate effect are within the analytical errors. Hence, the inference that the last prominent moraine-building event was quickly followed by deglaciation may not change in the future with refinements in the production rate.

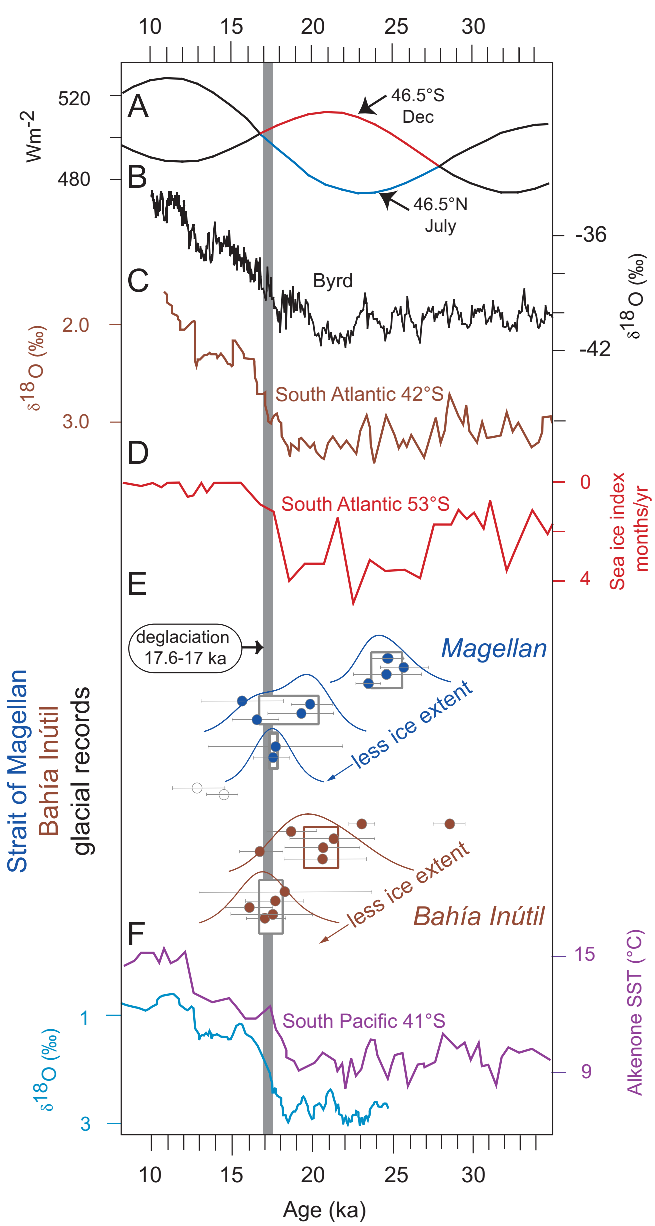
Only two ^{10}Be ages, 12.9 ± 1.7 and 14.5 ± 1.0 ka (SM-02-01 and 04), which are on the same recessional moraine crest (D) on the west side of the Strait, are statistically different than ^{14}C and stratigraphic data in the region (Heusser, 2003; McCulloch et al., 2005). Although it is not clear why the two ^{10}Be ages are younger than limiting ^{14}C data for this particular moraine crest, detailed mapping of the subdued landform and the surrounding area (Bentley et al., 2005) suggests the existence of ponded lakes during and after moraine formation, mobile loess, and trees before human settlement, all of which could have shielded erratics from cosmic radiation. This is the only site we collected from that was forested before human settlement (Bentley et al., 2005). Moreover, this moraine crest and the adjacent vicinity have been subject to intense animal grazing and this may have exposed the boulders in recent times. These two boulders are only 25–30 and 19–23 cm high, respectively. If covered by at least 3 m of water between deposition and end of the Lateglacial interval, or ~25 cm of overlying material was eroded by animals, the ages would appear about 5 ka too young.

Interestingly, this study demonstrates that in the Magellan region cosmogenic nuclide data on LGM deposits provide coherent ages for a given moraine limit or even individual crest (Table 2), which does not appear to be the case for pre-LGM deposits (Kaplan et al., 2007). For LGM moraines, ages are statistically indistinguishable on a given dated landform; e.g., for the D limit, reduced chi-squared (χ^2) values of <1 indicate that the samples are all from the 'same' population and that the analytical errors alone could explain the age scatter (Bevington and Robinson, 1992). Also, the ages in Table 2 follow within error the relative stratigraphic ages of the LGM moraines and the results are in agreement with ^{14}C assays (and are consistent with amino acid racimization data). In contrast, on pre-LGM deposits, generally the cosmogenic nuclide ages for a moraine limit or even a given landform do not all overlap at 1σ, they are from statistically different populations, or they are typically LGM in age. Furthermore, on the pre-LGM deposits, the cosmogenic nuclide data often violate other independent chronologic information. This finding led Kaplan et al. (2007) to conclude that geomorphic processes that inhibit cosmogenic dating of pre-LGM moraines may be less effective (i.e., within the analytical uncertainties) over the last ~25,000 years, specifically since deglaciation or during the Interglacial period.

The cosmogenic nuclide-based glacial chronology for the LGM in southernmost Patagonia shows similarities and differences with that obtained at Lago Buenos Aires, at 46°S (Kaplan et al., 2004; Douglass et al., 2006). Although the LGM is broadly similar between the two records there are also interesting differences (Kaplan and Moreno, in review). In both areas 4–5 moraine-building events are documented between ~25 and 16 ka. At LBA, the first preserved moraine-building event is ~23 ka, which is statistically indistinguishable from the 25 ka event in the Strait of Magellan. LBA ice is almost as extensive at ~15–14 ka as during late LGM time (Douglass et al., 2006), whereas the Straits were ice-free by ~17 ka.

6.2. Southern middle latitude paleoclimate

The Strait of Magellan-Bahía Inútil record provides unique terrestrial information at this latitude on glacial to interglacial ocean and climate changes between ~25 and 17 ka. Hulton et al. (2002) used a coupled climate/ice sheet model to reconstruct paleo-conditions that would lead to the growth and maintenance of a Patagonian ice sheet to the LGM mapped limit. The model calculates precipitation, which varies with prescribed temperature, windspeed and direction. Sensitivity experiments were run to study the growth and response of the ice sheet to different possible climatic forcings. They found that the modeled ice sheet was most sensitive to temperature and over southern Patagonia, or south of 43°S, it was necessary to reduce precipitation, otherwise ice overextended to the east. The 'best fit' to the empirical evidence was achieved with a 6 °C lowering.



A fundamentally different climatic regime existed between ~25 and 17 ka, marked by cooling at a time of high summer insolation intensity in the middle latitudes of the Southern Hemisphere. An explanation is that during this time there must have been prolonged northward migration and a strong influence of the Antarctic Frontal Zone, at present 3–5° to the south (where temperature drops considerably by 6 °C or more; Fig. 1b) (Belkin and Gordon, 1996; Olbers et al., 2004). The 'average' position of relatively low-pressure frontal systems surrounding Antarctica (Prohaska, 1976) may have migrated northward, associated with increased sea-ice extent (Fig. 4) (Stuut et al., 2004), or there may have been an increased frequency of mobile polar highs and, in general, dominance of 'cold' anticyclonic air masses (Leroux, 1993; Marengo and Rogers, 2001). A slight drying in southernmost Patagonia (Hulton et al., 2002), which is consistent with pollen-based reconstructions (Heusser, 2003), may have been due to either the core of the moisture-bearing westerlies moving northward, or a stronger orographic effect with the presence of a high ice sheet barrier (Sugden et al., 2002).

Other terrestrial studies in southernmost Patagonia, e.g., at >50°S, also have concluded that in the glacial world an expanded sea-ice edge and climate boundaries surrounding Antarctica strongly affected conditions on this part of the continent (Heusser, 1989, 2003; Markgraf et al., 1992). Based on fossil pollen data, Heusser (1989, 2003) and McCulloch et al. (2000) argued for northward migration of

Fig. 4. Records of southern latitude ocean-atmosphere response between ~30 and 10 ka, and for comparison insolation changes in the middle latitudes of both hemispheres. Colder is always downward. The Magellan region record is plotted between South Atlantic and southeastern Pacific records. (A) Insolation values for June at 45°N and S (Berger and Loutre, 1991). As these largely reflect changes in the intensity of summer insolation due to precession of the equinoxes, they are out-of-phase between the hemispheres. (B) Oxygen isotope record of the Byrd ice core (Blunier and Brook, 2001), Antarctica. (C) ^{18}O record from core RC11-83 measured on planktonic foraminifera from the SE Atlantic Ocean (Charles et al., 1996). Changes in planktonic δ^{18}O are inferred to reflect importance of subtropical or Sub Antarctic water masses and SST over the site (Fig. 1) (Charles et al., 1996; Lamy et al., 2004). (D) Sea-ice record from core TTN057-13 from the SE Atlantic Ocean at 53°S, about 4° south of the polar front and 3° north of the present winter sea edge (Stuut et al., 2004). (E) Timing of the LGM and major glacial advances in the Strait of Magellan (blue) and Bahía Inútil (light brown). All ^{10}Be ages are shown (from Table 2, erosion rate = 0 mm/ka and $\pm 1\sigma$ analytical uncertainty). For each moraine that has at least two measurements (5 total, Table 2), shown as rectangles are arithmetic means at ± 1 standard deviation and relative probability plots. For each respective lobe, towards the lower left the moraines represent less extensive ice margin extent (Fig. 2). The two youngest ages plotted from a single moraine crest in the southwestern Strait of Magellan (open white circles) are considered too young, as explained in text. Thick vertical line across figure represents timing of deglaciation of Strait of Magellan–Bahía Inútil, which the ^{14}C data indicate was complete by 17.6–17.0 cal ka (Figs. 2 and 3) (Heusser, 2003; McCulloch et al., 2005). (F) Alkenone temperature and planktonic foraminifer δ^{18}O records from core ODP1233 in the southeast Pacific Ocean (Lamy et al., 2004).

the core of the westerlies; whereas, Markgraf (1989) and Markgraf et al. (1992) discussed that the band of (relatively) fastest flow intensified and was displacement southward of the continent, consistent with some GCM modeling studies (Wyrwoll et al., 2000; Whitlock et al., 2001). Although our data do not directly address this important debate, as in this study, these researchers argued for increased influence of Polar air on the southern South American continent. Ackert et al. (2003) also inferred northward movement of air and climate boundaries and the Antarctic convergence during the Last Glacial cycle to explain apparent low atmospheric pressure over southern Patagonia based on higher than expected concentrations of in situ cosmogenic ^{3}He in ^{40}Ar/^{39}Ar dated lava flows.

Comparison of the southern South American record with those from the neighboring South Pacific and Atlantic oceans, and in Antarctica, documents a similar history of environmental conditions (Figs. 1 and 4), even considering that present uncertainties in ^{10}Be ages prevent precise correlation at the millennial scale between moraine-building events and ocean records. Specifically, sea ice, alkenone sea surface temperature, and planktonic δ^{18}O all indicate 'glacial conditions' between ~28 and 17 ka, with a temperature reduction relative to modern of ~6 °C (Fig. 4). Maximum sea-ice extent in the South Atlantic Ocean occurred around 27–22 and 21–18 ka, overlapping in time with the B–D major moraine-building periods. Although the glacial record 'begins' at the maximum Patagonian ice extent, ~25 ka, prior glacial activity during an earlier part of the terrestrial LGM could have been removed. Relatively limited net retreat between ~25 and 20 ka, or during the formation of the 'B and C' limits, on the Juan Mazia Peninsula area may mean that the glacier margin remained close to the maximum extent over the span of cosmogenic ages (within 1σ) or that it fluctuated and readvanced to a similar position, while Southern Ocean conditions remained cold. Slight warming is indicated in marine proxies and in Antarctica before 18 ka; similarly, the D limit is well within the B–C limits of middle and early LGM time, reflecting limited, but some net ice retreat and surface lowering before 18 ka (Benn and Clapperton, 2000). Subsequently, deglaciation of southernmost Patagonia was well underway by 17 ka, consistent with that in Southern Ocean and Antarctic records (Fig. 4). Although the ice sheet may have responded to an abrupt shift in climate regime after 18 ka, it is also possible that the grounding line was unstable after retreat from Juan Mazia causing rapid deglaciation of the area (Porter et al., 1992).

Collectively, the southern latitude marine and terrestrial records imply at least 3–5° of 'mean' northward movement of cold ocean–air masses around the Drake Passage and southern South America (Fig. 1) between ~25 and 17 ka. Although the present positions of the oceanic fronts around the Drake Passage appear to be fixed by bottom topography, the marine evidence strongly supports an equatorward shift in environmental surface conditions in both the neighboring South Pacific and Atlantic oceans during the LGM (e.g., Anderson et al., 2002). Of note is that δ^{18}O data in the Byrd ice core and in marine cores were much closer to glacial than interglacial values during the latter part of isotope stage 3, well before the Patagonian LGM apparently began around 25 ka. By implication, it can be hypothesized that late stage 3 glaciers may have existed in the Andes at a larger size than those at present and they expanded relatively quickly around 25 ka (or slightly before), in phase with the northward migration of increased sea ice. For comparison, all proxies show change towards interglacial values around or after ~18 ka.

7. Conclusions

Southern Patagonian ice was most extensive from ~25 to ~18 ka, with a peak around 25–24 ka. This was a period of rising and maximum summer insolation intensity in the Southern Hemisphere largely due to the precession of the equinoxes (Fig. 4). A stronger influence of the Antarctic Frontal Zone and equatorward movement of cold air from the south relative to the present is one way to explain negated local summer insolation intensity, and a glaciation broadly in phase with global changes. Terrestrial and marine findings for the Southern Ocean are supported by Patagonian modeling results, which all indicate expansion of the polar climate regime, influencing atmosphere–ocean conditions and southern Andes' ice age history. The climatic setting at the start and end of the LGM may have been different, perhaps with conditions primed for terrestrial glaciation before, but not after.

The terrestrial–marine paleoclimate history in the southern latitudes can be used to address questions of interhemispheric phasing of climate change during the Last Glacial period and Termination I (Broecker and van Donk, 1970). The time of extensive ice in the Magellan region is broadly similar to the middle latitude glacial history of the Laurentide Ice Sheet (Dyke et al., 2002), and considering dating uncertainties, is only slightly before or overlapping in time with the global LGM (Mix et al., 2001). In addition, the specific period of most extensive glacial ice, ~25–24 ka, was coeval with the coldest temperatures around the North Atlantic region (Alley et al., 2002). On the other hand, the terrestrial–marine data suggest that the Termination of the Last Glacial cycle was well underway by 17 ka in the mid-high southern latitudes, which bears similarities with other terrestrial records in both hemispheres, but was noticeably before major change in the North Atlantic region (McManus et al., 2004; Schaefer et al., 2006).

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